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# Thermocline of the Flores and Banda Seas

Arnold L. Gordon and Amy Ffield[1]

Lamont-Doherty Earth Observatory, Palisades, New York

A. Gani Ilahude

Pusat Penelitian Dan Pengembangan Oseanologi, Indonesian Institute of Sciences, Jakarta, Indonesia

**Abstract.** In December 1991, 30 conductivity-temperature-depth (CTD) stations to 1000 dbar were obtained from the R/V *Baruna Jaya I* in the Flores Sea, Banda Sea, and Alor-Wetar passage of the Indonesian sea. A salinity maximum within the interval 100–150 dbar and a salinity minimum within the interval 300–350 dbar mark water mass core layers derived from the North Pacific. They are drawn into the Flores Sea from the Makassar Strait, with subsequent flow into the Banda Sea, and are weakened en route by strong vertical mixing characteristic of the Indonesian seas. In the Flores Sea, water below 300 dbar becomes saltier with increased distance from the Makassar Strait, suggesting that an advective process may be drawing relatively salty water into the Flores Sea lower thermocline from the Banda Sea. The Banda-to-Flores Sea flow may be a consequence of vertical transfer of horizontal momentum produced by the same turbulent processes that are responsible for enhanced vertical mixing. The interocean transport profile may not correspond exactly with the Pacific-to-Indian Ocean pressure gradient profile, as deeper water is carried along with the through flow by the effects of eddy viscosity. The 550-m sill at the southern end of the Makassar Strait creates a situation where downward flux of momentum entrains deeper water that must be compensated by lower thermocline water drawn from the Banda Sea. Geostrophic transport relative to 1000 dbar in the Banda Sea shows not only a strong through flow transport in the upper 300 dbar ($6.3 \times 10^6$ m$^3$ s$^{-1}$) but also a deeper flow toward the Flores Sea ($1.5 \times 10^6$ m$^3$ s$^{-1}$ from 300 to 500 dbar and an additional $2.4 \times 10^6$ m$^3$ s$^{-1}$ from 500 to 1000 dbar). A simple model suggests that the magnitude of the deeper westward flow is proportional to the vertical eddy viscosity coefficient. Water mass analysis shows that either the South Pacific or Indian Ocean can provide the lower thermocline Banda Sea water.

## 1. Introduction

The Indonesian seas are a conduit for interocean flow of tropical thermocline water from the Pacific Ocean to the Indian Ocean [*Wyrtki*, 1961; *Gordon*, 1986]. There is much attention in recent literature to this through flow [e.g., *Hirst and Godfrey*, 1993; *Inoue and Welsh*, 1993]. While this transfer has an important large-scale influence [*Gordon*, 1986; *Toole et al.*, 1988], the effect of enhanced vertical mixing characteristic of the Indonesian seas [*Ffield and Gordon*, 1992] must also be considered, in that the associated large vertical fluxes significantly alter the stratification of the through flow water and, as is shown in this paper, may also influence the velocity profile. The turbulent processes (processes presumably induced by the strong tidal currents characteristic of the Indonesian seas) that are responsible for enhanced vertical eddy mixing also would produce enhanced vertical flux of horizontal momentum.

The interocean through flow is driven by the Pacific-to-Indian Ocean pressure gradient, confined for the most part to the upper 200 dbar, with strong seasonal and interannual variability [*Wyrtki*, 1987; *Kindle et al.*, 1989]. The through flow is high during the southeast monsoon (June to August) and low during the northwest monsoon (December to February) [*Wyrtki*, 1987]. Water mass analysis [*Ffield and Gordon*, 1992] shows that most of the thermocline water is derived from the North Pacific Ocean, with the main pathway through Makassar Strait. Makassar Strait water enters the Indian Ocean via the Lombok Strait [*Murray et al.*, 1989], though a substantial part turns eastward into first the Flores Sea and then the Banda Sea.

The surface layer is influenced by local wind forcing (as discussed by *Wyrtki* [1958, 1961], *Birowo and Ilahude* [1977], *Zijlstra et al.* [1990], and *Ilahude et al.* [1990]). During the southeast monsoon, from June to August, surface water is driven from the Banda Sea into the Flores, Jawa, and South China Seas. During the northwest monsoon, from December to February, surface water from the Java (Jawa) Sea and Makassar Strait is driven across the Flores Sea into the Banda Sea.

In December 1991 a cruise aboard the Indonesian R/V *Baruna Jaya I* (BJI) in the Flores and Banda Seas and within the Alor-Wetar passage allowed inspection of the thermohaline stratification and associated circulation pattern of this region. Thirty conductivity-temperature-depth (CTD) stations to 1000-dbar pressure were obtained as part of the Irja

[1]Now at Cooperative Institute for Marine and Atmospheric Studies, Miami, Florida.



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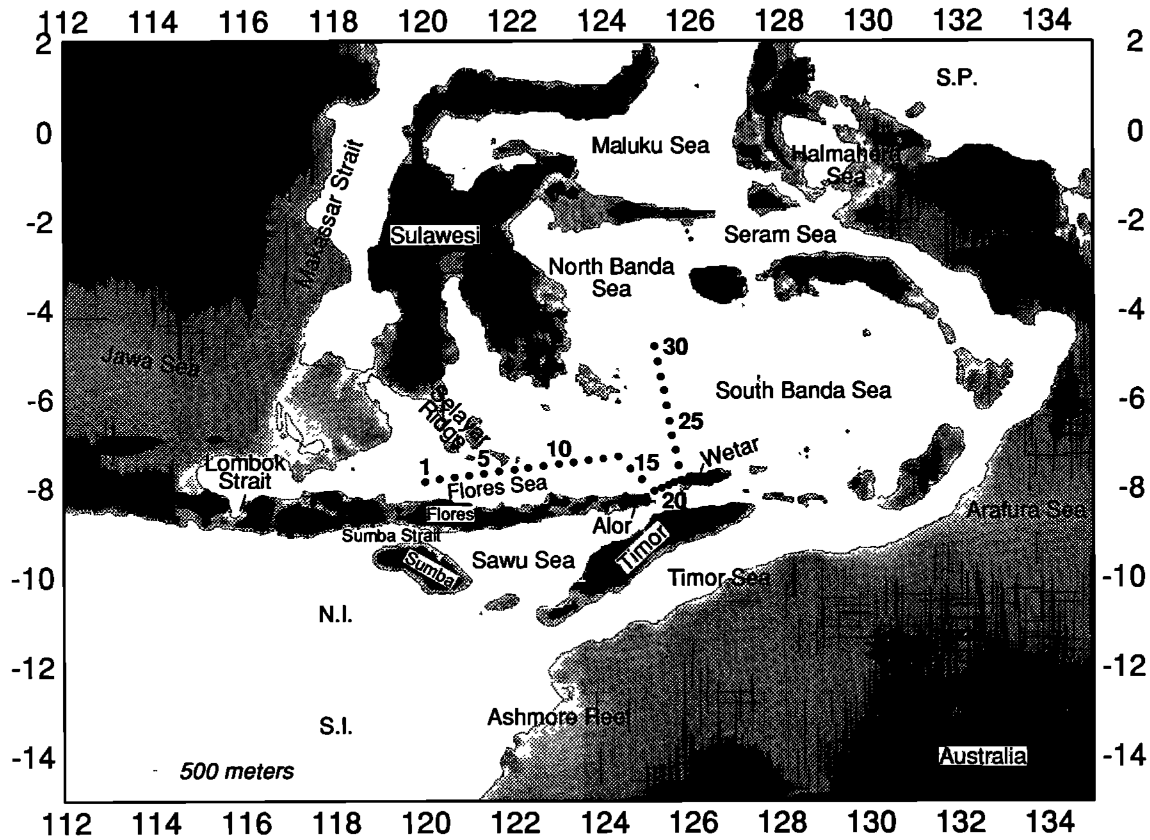


**Figure 1.** R/V *Baruna Jaya I* conductivity-temperature-depth stations 1–30 obtained in December 1991. N.I. refers to north Indian Ocean; S.I., to south Indian Ocean; and S.P., to South Pacific Ocean.

Zee–91 expedition (Figure 1). In December the northwest monsoon should have already initiated heavy rain and strong west wind in the region, but in 1991 it was delayed by an El Niño–Southern Oscillation event that persisted for much of 1991 [*Darwin Tropical Diagnostic Statement* (*DTDS*), 1991]. The DTDS report also states that light wind induced a +1°C temperature anomaly in December 1991 resulting in a sea surface temperature greater than 30°C within the southern Makassar, Flores, and Banda Seas.

In this paper the BJI data are used to study the evolution of the upper 1 km of the water column along the flow path from Makassar Strait across the Flores Sea into the Banda Sea. A concept is introduced suggesting that the influence of strong vertical flux of horizontal momentum within the Indonesian seas accelerates water residing deeper than the interocean pressure gradient characteristic of the upper 200 dbar [*Wyrtki*, 1987]. For example, where the flow from the Pacific Ocean to the Indian Ocean encounters a sill, as in the southern Makassar Strait, a circulation pattern may be established to the lee of the sill, with deeper water flowing toward the sill and upwelling to join the Pacific-to-Indian transport. We first present the basic stratification, and then in the discussion section we draw this information together to support the circulation concept for the Flores and Banda Seas.

## 2. Station Data

A BJI station consists of a Guildline CTD lowered to approximately 1000 dbar, tripping 12 water samplers for salinity and oxygen determination (Figures 2a and 2b). Station 30, the only station with a full set of bottle salinities (Figure 2b), indicates that the difference between bottle salinity and CTD salinity is about 0.015 practical salinity units (psu). The CTD temperature error is estimated as ±0.01°C. In the following analysis the uncorrected CTD values are used, since the errors are too slight to influence the figures or results discussed in this paper.

## 3. Stratification

### 3.1. Flores Sea and Western Banda Sea (Stations 1–16 and 23)

The Flores Sea reveals a more structured salinity profile (Figure 2a) in comparison to the Banda Sea (Figure 2b). This reflects the proximity of station 1 (Figure 2a) to Makassar Strait, the primary through-flow pathway. Water mass analysis [*Wyrtki*, 1961; *Ffield and Gordon*, 1992] traces both the salinity maximum (s-max) near 110 dbar (20°C) and the salinity minimum (s-min) near 300 dbar (10°C) to the North Pacific Subtropical Lower Water and Intermediate Water, respectively. With surface water approximately 29°–30°C and the 10°C isotherm falling near 325 dbar, the Flores Sea thermocline is intense.

In both the Flores and Banda Seas the surface salinity is markedly fresher than that of the subsurface water (Figures 3a and 3b). This is a consequence of accumulation of regional river discharge and rainfall. The lowest surface salinity occurs in the western Flores Sea (Figures 4a and 4b;

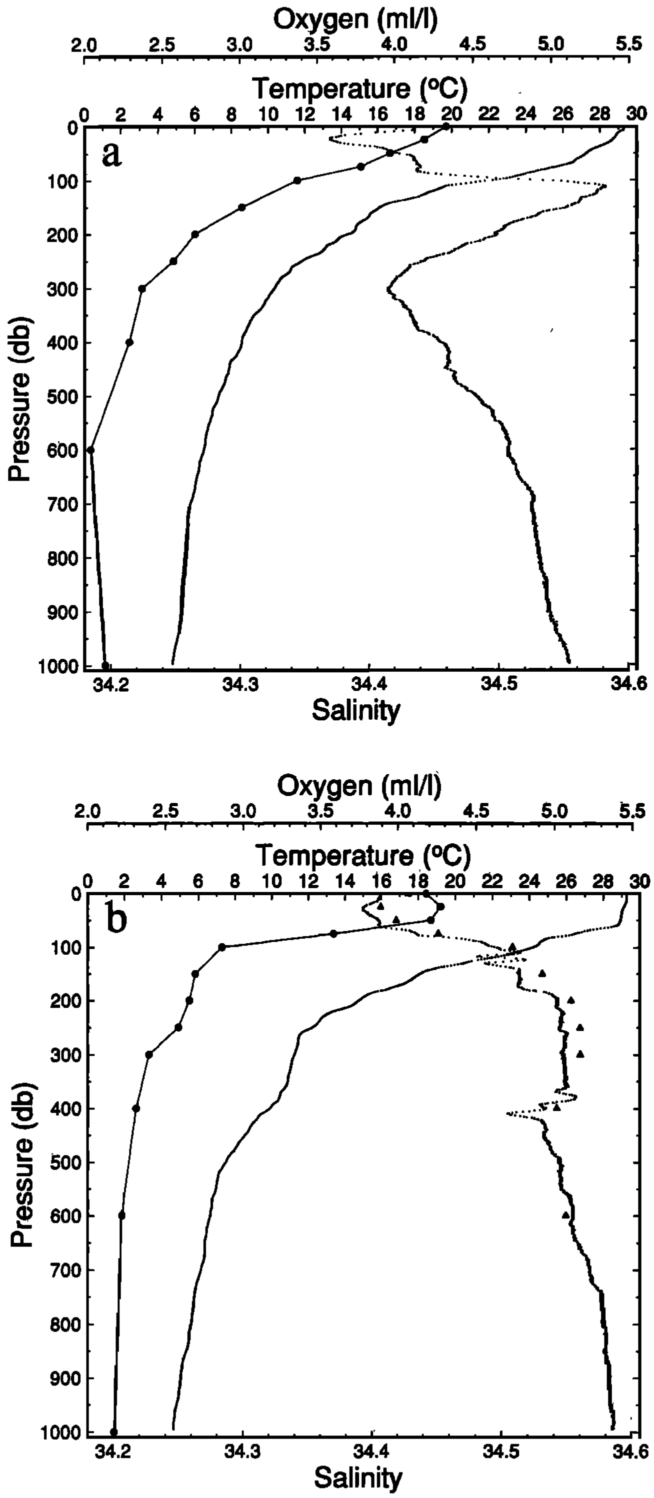


**Figure 2.** Temperature, salinity, and oxygen profiles for (a) station 1 and (b) station 30. Salinity values obtained from sample bottles are shown as triangles, and oxygen values are shown as dots.

Figure 4a is included for completeness), some of which may be supplied by outflow from the Java Sea. Surface salinity approaches 34.5 in the eastern Flores Sea and remains above 34.4 within the Banda Sea.

Within the western Banda Sea (stations 10–14 and 23), both salinity extrema are greatly attenuated, producing a nearly isohaline thermocline (Figure 4b). The separation of the Flores and Banda Seas occurs rather abruptly in the region of the Selayar Ridge (Figure 1) and surrounding islands and coral reefs, as well as in deep passages. Within this transition (stations 5–9) the stratification becomes more complex, as the s-max is composed of multifilaments. It is likely that this stratification is produced as Flores and Banda thermocline water, each with somewhat different mixing histories, are brought together by circulation patterns asso-

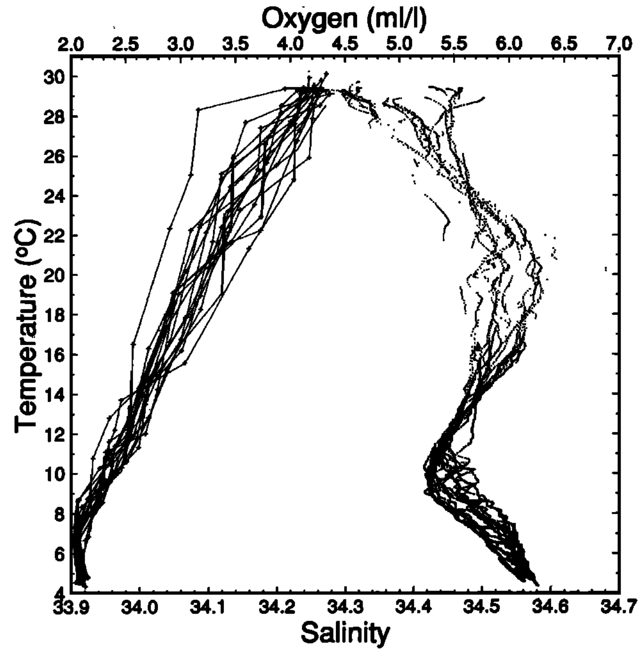


**Figure 3a.** Temperature-salinity and temperature-oxygen relationships for stations 1–16 of the Flores Sea section.

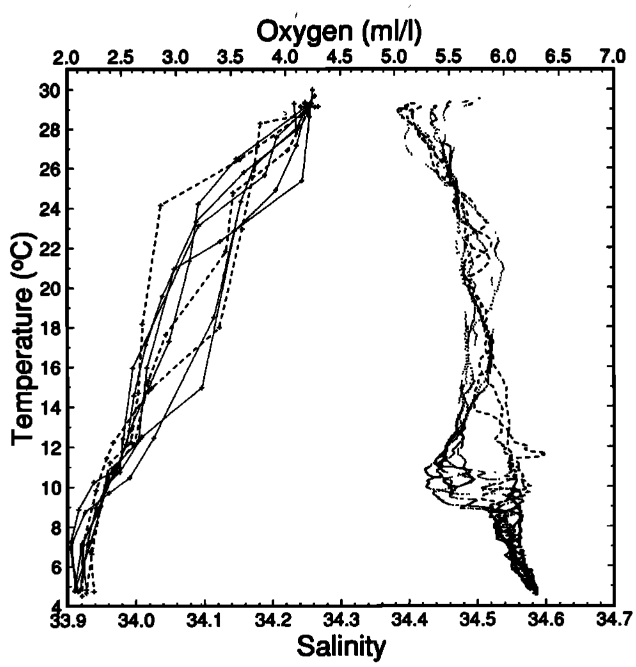


**Figure 3b.** Temperature-salinity and temperature-oxygen relationships for stations 22–30 of the Banda Sea section. Dashed lines represent stations 28, 29, and 30.

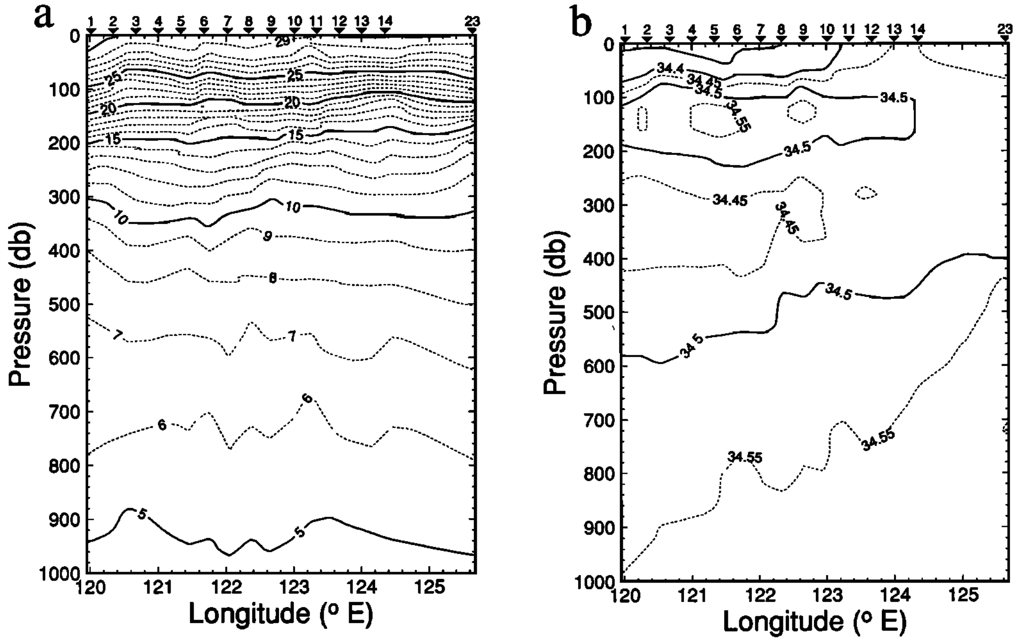


**Figure 4.** Flores Sea section using stations 1–14 and 23 for (a) temperature and (b) salinity.

ciated with the Selayar Ridge. The Banda Sea weaker s-max water may be introduced to the region by a route along the Sulawesi coast that turns to the south at the Selayar Ridge. As vertical mixing is responsible for the s-max destruction, the interleaving pattern that increases the mean vertical gradient would be expected to accelerate s-max destruction.

### 3.2. Banda Sea (Stations 21–30)

Below 100 dbar the Banda Sea (Figure 2b) displays a nearly isohaline structure relative to the Flores Sea (Figure 2a). The deep s-max at 100–200 dbar and the deep s-min at 250–350 dbar dominating the southern Banda Sea between stations 22 and 27 (Figures 3b, 5a, and 5b; Figure 5a is included for completeness) are similar to the Flores Sea stratification, though with more advanced destruction of the extrema, as expected from its downstream position. The s-max values are reduced to less than 34.55 for all of the Banda Sea stations. The s-min layer of the southern Banda Sea survives for water warmer than 10°C but is essentially removed for water colder than 10°C. In stations 28–30 of the northern Banda Sea, the s-min layer is absent even for water warmer than 10°C.

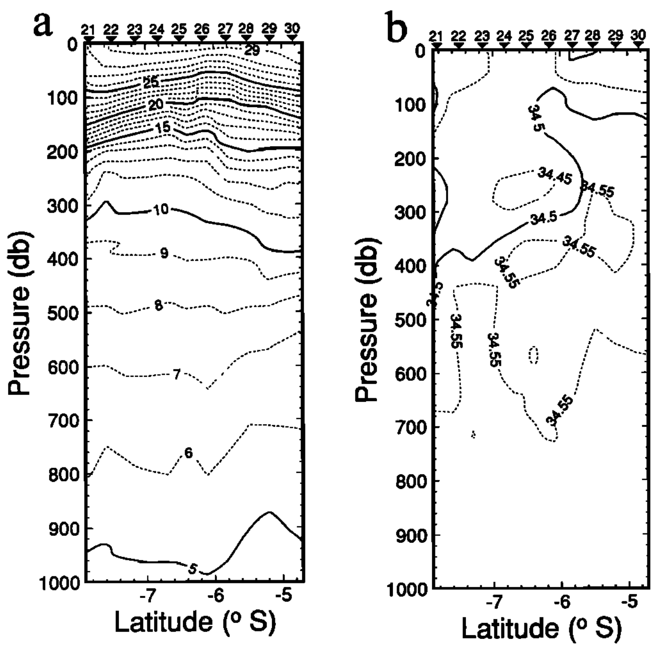


**Figure 5.** Banda Sea section using stations 21–30 for (a) temperature and (b) salinity.

### 3.3. Salt Content

The transition of the Flores Sea profile to that of the Banda Sea may at first be considered to be due solely to vertical mixing effects. In order to evaluate this hypothesis, the change in salt content of the water column relative to station 1 is plotted for the upper 300 dbar and the intervals 300–500 dbar and 500–1000 dbar (Figure 6). The salt content of the upper 300 dbar remains nearly constant (±5 kg m$^{-2}$), consistent with a vertical mixing process that simply redistributes the salt (Figure 6). However, the salt content of the water in the 300- to 1000-dbar interval increases with distance from the Makassar Strait. The salt buildup is too great to be supplied by vertical processes across the 1000-dbar level. Multiplying the salinity gradient at 1000 dbar with a vertical mixing coefficient $K_z$ of $1 \times 10^{-4}$ to $10 \times 10^{-4}$ m$^2$ s$^{-1}$ units supplies 0.3–3 kg m$^{-2}$ yr$^{-1}$, into the upper 1000 dbar. As the total salt increase for the upper 1000 dbar is calculated as 45–50 kg m$^{-2}$, the thermocline would require a residence time of 150–15 years for the 1–10 span in $K_z$. Thermocline residence time is likely to be less than this [*Ffield and Gordon*, 1992]. It is proposed that the salt increase in the 300- to 1000-dbar interval is due to lateral injection of salty lower thermocline water into the Banda Sea, with subsequent spreading into the Flores Sea. Sources of the salty water are discussed in section 5.2.

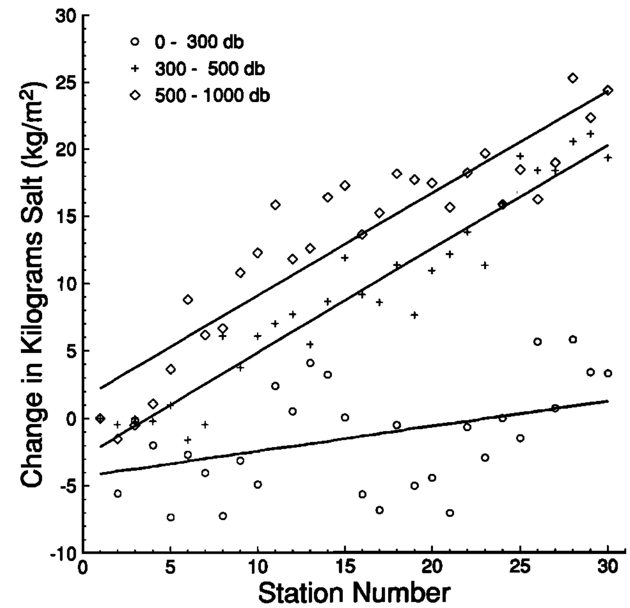


**Figure 6.** Change in salt content (in kilograms salt per square meter) for the intervals 0–300, 300–500, and 500–1000 dbar for each station relative to station 1 with a linear fit.

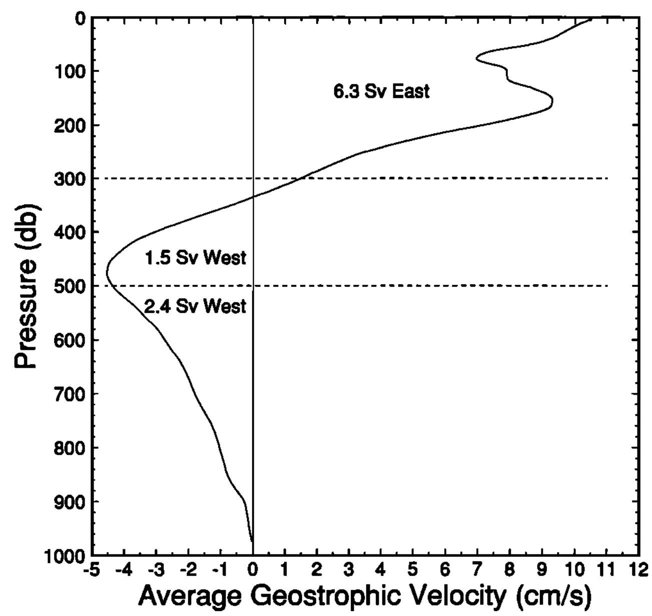


**Figure 7.** Geostrophic velocity relative to 1000 dbar between stations 22 and 30 of the Banda Sea. The transport within the 0- to 300-dbar, 300- to 500-dbar, and 500- to 1000-dbar levels is shown in sverdrups (1 Sv = $10^6$ $m^3$ $s^{-1}$).

### 3.4. Oxygen Stratification

The oxygen stratification for station 1 (Figure 2a) and station 30 (Figure 2b) shows decreasing oxygen with increasing depth. However, the majority of stations show somewhat higher oxygen values near 250–300 dbar relative to a simple exponential oxygen versus depth trend (Figures 3a and 3b), generally near the 12°C isotherm (this is best seen in the station 30 profile, Figure 2b); conversely, one may say the oxygen is relatively low near the 7°C isotherm. We suggest that since there is no source of relatively oxygenated water near the 12°C isotherm [*Wyrtki*, 1961], the relatively low-oxygen water near the 7°C isotherm may be drawn from the Indian Ocean, specifically from a Persian Gulf influence through the Alor-Wetar passage [*Ilahude*, 1992a, b].

## 4. Geostrophic Transport

During the transition from the southeast monsoon to the northwest monsoon, the surface flow changes direction with the wind. The subsurface water that responds to the interocean pressure gradient may remain in more of a steady balance. In order to relate the velocity field with the water mass distributions, geostrophic transport relative to 1000 dbar is presented (Figure 7) for the same layers used for the salt content calculation (Figure 6). In the Banda Sea the net upper layer flow is eastward at 6.3 Sv (1 Sv = $10^6$ $m^3$ $s^{-1}$); from 300 to 1000 dbar the flow is westward at 3.9 Sv. The 300- to 1000-dbar layer provides deep inflow of relatively salty water to the Flores Sea (Figure 6). Geostrophic calculations at low latitudes are fraught with errors (internal wave and tidal displacement of isopycnals); hence they cannot be taken as primary evidence for a phenomenon and must be considered only in a qualitative sense. In defense of the geostrophic transport values (Figure 7) we note that stations 22 and 30 used for the calculations are 23 hours and 32 min apart, making the tidal oscillations a minor factor. Additionally, as these stations are 300 km apart, the effects of internal waves are also minimized because the geostrophic tilt is large over that distance. Also, when we used other greatly separated station pairs in the Banda Sea, we obtained similar transport profiles.

## 5. Discussion

### 5.1. Circulation of the Flores Sea

The Pacific-to-Indian Ocean pressure gradient of the upper 200 dbar drives the interocean transport within the Indonesian seas with a retardation by the frictional effects [*Wyrtki*, 1987; *Inoue and Welsh*, 1993]. As the Indonesian seas' thermocline has enhanced vertical mixing [*Ffield and Gordon*, 1992], it follows that the interocean through flow momentum derived from the upper 200 dbar pressure head would be transferred to deeper water by eddy viscosity. The interocean transport profile may not correspond exactly with the pressure head profile, in that deeper water would be "dragged" toward the Indian Ocean. The interocean flow of North Pacific Intermediate Water may benefit from this momentum flux. At some point the downward flux of momentum would pass below the sill of the southern Makassar Strait (Figure 1). To the lee of the sill, subsill water is "dragged" toward the Indian Ocean by the Pacific-to-Indian Ocean flow. The subsill entrained water must be replenished from still deeper water drawn from the downstream direction (Figure 8).

Analysis of the water mass stratification of the Flores and Banda Seas supports this circulation pattern concept. In the Banda Sea the upper 300 dbar has about the same salt content as the Flores and hence is North Pacific in origin, but below 300 dbar the salt content is well above that of the Flores Sea and cannot be drawn from the North Pacific via the Flores Sea. The salt enhancement of the Banda Sea is carried into the lower thermocline water of the Flores Sea by the circulation induced below the Makassar sill.

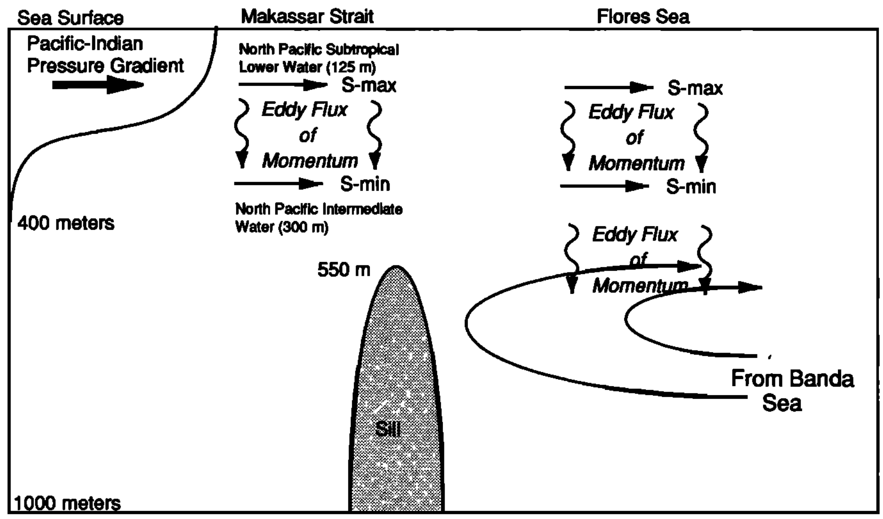


**Figure 8.** Schematic of the circulation pattern proposed for the Flores Sea. Continuous downward eddy flux of momentum to depths below the Makassar Strait–Flores Sea sill depth drives a deeper inflow of salty water from the Banda Sea.

A simple model is constructed to estimate the structure and potential vigor of the Flores Sea vertical plane thermocline circulation as a function of the vertical eddy viscosity coefficient $A_z$ [*Ffield*, 1994]. Conceptually, there are three layers: layer 1, the upper 200 dbar through-flow layer with a constant horizontal through-flow velocity; layer 2, where horizontal pressure is assumed to balance vertical eddy viscosity

$$\delta P/\delta x = \rho A_z \, \delta^2 u/\delta z^2$$

between 200 dbar and the depth of the viscosity influence; and layer 3, the deep layer, where the velocity is taken as zero. It is assumed that the interocean pressure head maintains the momentum of the upper 200 dbar and that momentum from the upper through-flow layer is transferred to the otherwise motionless layer 2 by eddy viscosity. This momentum flux induces a horizontal flow in layer 2 in the same direction as the through flow. Because a sill restricts horizontal flow in layer 2 from upstream, the induced subsill flow must be drawn from below and downstream where horizontal access to deeper Banda Sea water is possible. This is modeled by requiring that the net horizontal flow in layer 2 equal zero.

In order to demonstrate the concept most simply, a quadratic form of the horizontal velocity profile is taken for layer 2 and solved using layers 1 and 3 for upper and lower boundary conditions, respectively, and requiring that the net horizontal flow in layer 2 equal zero. The resulting velocity equation and a pressure equation found by utilizing simple expressions for sea surface height gradient, interface gradient, and layer densities are substituted into the pressure-viscosity equation. The depth of the viscosity influence can be solved for and then used with the velocity profile to estimate the model layer transport.

A north-south pressure gradient in the Flores Sea would geostrophically balance the west-east induced flow of the model. Therefore the model velocities are compared with the west-east geostrophic transport profile calculated from Banda Sea stations 22 and 30 (Figure 7). The similarity in shape and magnitude of the calculated and modeled velocity profiles demonstrates the potential validity of our simple model (Figure 9a). The transport and depth of the model layer increases by the square root of the vertical eddy viscosity coefficient (Figure 9b). With an $A_z$ range from $1 \times 10^{-4}$ m$^2$ s$^{-1}$ to $1 \times 10^{-3}$ m$^2$ s$^{-1}$, the model layer inflow to the Flores Sea varies from 0.4 to 1 Sv (Figure 9b), and the model layer depth varies from 90 to 270 m.

### 5.2. Source of Lower Thermocline Water

Comparison of the Banda Sea salty lower thermocline water with that of the surrounding oceans indicates that the South Pacific Ocean or Indian Ocean could provide the source of the relatively salty lower thermocline of the Banda Sea (Figures 10a and 10b). A South Pacific source would be drawn into the Indonesian seas through the Molucca (Maluku) and Halmahera Seas. Indian Ocean water may be

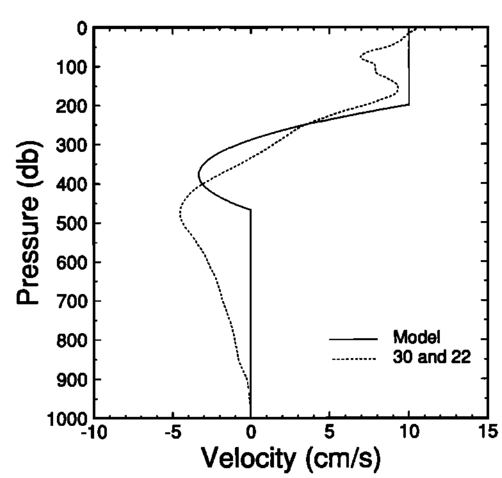


**Figure 9a.** Model-predicted horizontal velocity profile compared with geostrophic velocity profile calculated relative to 1000 dbar between stations 22 and 30 of the Banda Sea.

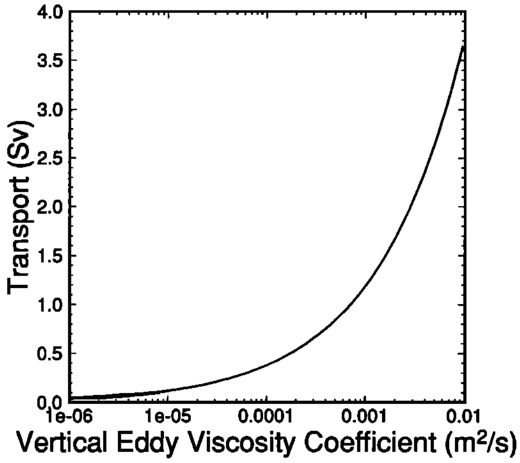


**Figure 9b.** Transport of the lower layer inflow to the Flores Sea as a function of vertical eddy viscosity coefficient $A_z$.

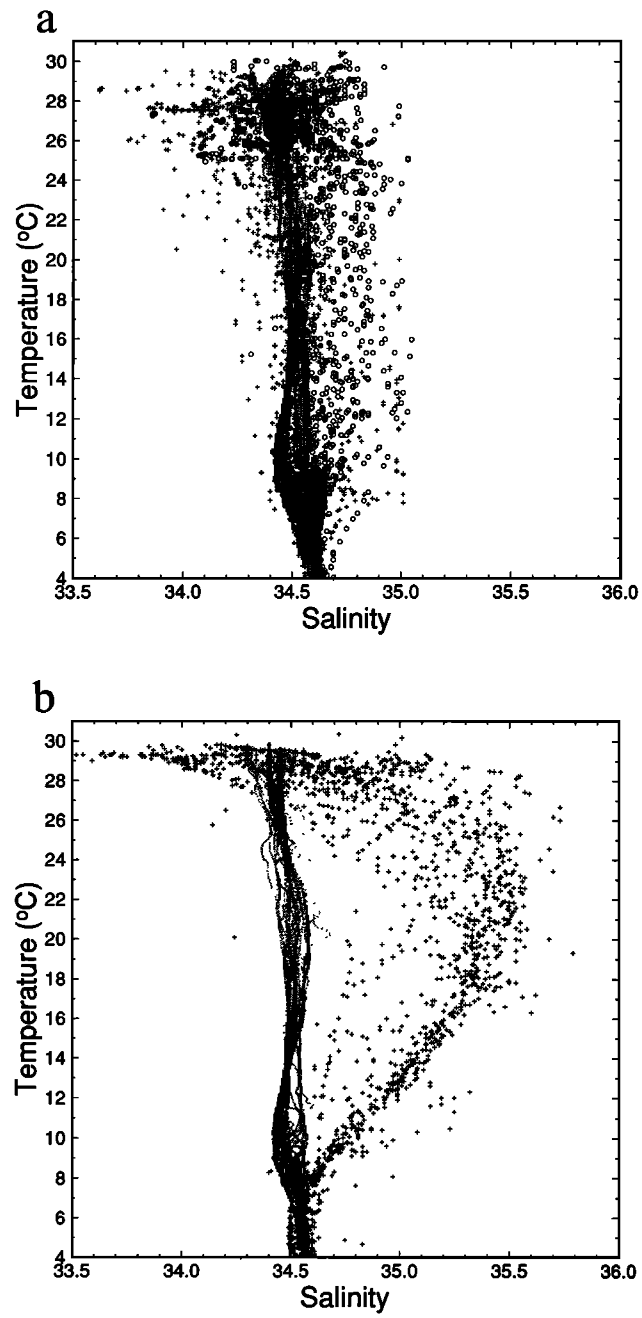


**Figure 10.** Temperature and salinity plotted for the Banda Sea (a) with South Pacific archived data and (b) with Indian Ocean archived data. In Figure 10a, crosses represent South Pacific Ocean water. In Figure 10b, crosses represent north Indian Ocean water, and circles represent south Indian Ocean water south of the strong thermohaline front near 12°S [*Wyrtki*, 1971]. See Figure 1 for the general location of the archived data.

derived from two sources: South Indian Ocean thermocline water or Persian Gulf Intermediate Water (PGIW) derived from the northern Indian Ocean [*Wyrtki*, 1961]. An Indian Ocean source has access to the Banda Sea via the Alor-Wetar passage and the passages on either side of Timor.

South Indian Ocean water migrates eastward along the margin of northwest Australia in the Timor Sea at 300 dbar [*Wyrtki*, 1961, Plate 34]. The northern edge of the Timor Sea has lower-salinity water much like that of the Banda Sea. While most of the South Indian Ocean water may be turned back to the west by the Ashmore reef of the Australian margin, it is possible that some Indian Ocean water, particularly at depths greater than 1000 dbar, may continue to spread to the east and subsequently into the eastern Indonesian seas (as found through Pegasus measurements by *Fieux et al.* [1994]). A South Indian source would flow along the eastern deep passages between the South Banda and Arafura Seas, entering the Ceram (Seram) Sea from the east and then flowing into the Banda Sea to the south (a pattern for deep water discussed by *Van Bennekom et al.* [1988]). Along this rather long route continuous upwelling would allow deeper Indian Ocean water to eventually reach the intermediate and lower thermocline stratum discussed in this paper.

A strong eastward subsurface flow of saline Persian Gulf Intermediate Water has been inferred from water mass analysis and geostrophic flow along the southern coast of Java that passes through the Sumba Strait (between Flores and Sumba Islands) into the Savu Sea but not back to the south with the through flow between Timor and Sumba [*Ilahude*, 1992a, b]. The PGIW, with a temperature between 8° and 10°C and salinity from 34.65 to 34.70 and marked by an oxygen minimum of about 2.0 mL $L^{-1}$, has the right properties to supply the saltier lower thermocline water of the Banda Sea. The BJI data suggest that the PGIW flows into the southern Banda Sea via the Alor-Wetar passage, in that oxygen levels in the interval 8°–10°C are below that of the Banda Sea in the vicinity of the Alor-Wetar passage.

An Indian Ocean source for the salty lower thermocline water would not contribute to interocean through flow; that is, the source does not spread into the Pacific Ocean but rather gets flushed back to the Indian Ocean within the lower thermocline.

## 6. Conclusions

Transport of North Pacific water into the Indian Ocean is forced by a interocean pressure head for the upper 200 dbar. Strong, turbulent processes within the Indonesian seas, most likely derived from tidal current interaction with the complex interconnecting array of basins, straits, shelves, and ridges, drives large vertical fluxes of ocean properties, including heat, salinity, nutrients, and momentum. The downward flux of momentum carries the effects of the pressure head to deeper layers, involving these deeper layers in the interocean through flow. The involvement of deeper water causes the net thermohaline flux to be altered as deeper, colder water joins the upper layer through flow. Once the 550-m controlling sill of the southern end of Makassar Strait is crossed, continued downward flux of momentum sets up an estuary type circulation in the Flores Sea. Relatively salty lower thermocline water (approximately 7°–8°C) from the Banda Sea is injected into the Flores Sea, where it upwells and is carried back to the east. It is estimated that the quantity of water drawn into the Flores Sea from the Banda Sea is of the order of a few sverdrups. The source of the salty Banda Sea water can be from either the South Pacific Ocean or the Indian Ocean.

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A. Ffield, Cooperative Institute for Marine and Atmospheric Studies, 4600 Rickenbacker Causeway, Miami, FL 33149. (e-mail: ffield@aoml.erl.gov)

A. L. Gordon, Lamont-Doherty Earth Observatory, Palisades, NY 10964. (e-mail: a.gordon@ldeo.columbia.edu)

A. G. Ilahude, Pusat Penelitian Dan Pengembangan Oseanologi, Jl. Pasir Putih 1, Ancol Timur, P.O. Box 4801-JKTF, Jakarta 11048, Indonesia.

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